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Future changes in precipitation and impacts on extreme streamflow over Amazonian sub-basins

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Abstract
Because of climate change, much attention is drawn to the Amazon River basin, whose hydrology has already been strongly affected by extreme events during the past 20 years. Hydrological annual extreme variations (i.e. low/high flows) associated with precipitation (and evapotranspiration) changes are investigated over the Amazon River sub-basins using the land surface model ORCHIDEE and a multimodel approach. Climate change scenarios from up to eight AR4 Global Climate Models based on three emission scenarios were used to build future hydrological projections in the region, for two periods of the 21st century. For the middle of the century under the SRESA1B scenario, no change is found in high flow on the main stem of the Amazon River (Óbidos station), but a systematic discharge decrease is simulated during the recession period, leading to a 10% low-flow decrease. Contrasting discharge variations are pointed out depending on the location in the basin. In the western upper part of the basin, which undergoes an annual persistent increase in precipitation, high flow shows a 7% relative increase for the middle of the 21st century and the signal is enhanced for the end of the century (12%). By contrast, simulated precipitation decreases during the dry seasons over the southern, eastern and northern parts of the basin lead to significant low-flow decrease at several stations, especially in the Xingu River, where it reaches −50%, associated with a 9%

reduction in the runoff coefficient. A 18% high-flow decrease is also found in this river. In the north, the low-flow decrease becomes higher toward the east: a 55% significant decrease in the eastern Branco River is associated with a 13% reduction in the runoff coefficient. The estimation of the streamflow elasticity to precipitation indicates that southern sub-basins (except for the mountainous Beni River), that have low runoff coefficients, will become more responsive to precipitation change (with a 5 to near 35% increase in elasticity) than the western sub-basins, experiencing high runoff coefficient and no change in streamflow elasticity to precipitation. These projections raise important issues for populations living near the rivers whose activity is regulated by the present annual cycle of waters. The question of their adaptability has already arisen.

Keywords: Amazon, ORCHIDEE, streamflow extreme, climate change, precipitation

S Online supplementary data available from stacks.iop.org/ERL/8/014035/mmedia

1. Introduction

Many questions arise about the impact of climate change on water resources, which are fundamental for ecosystems and societies. Such questions are especially relevant in the Amazon River basin where severe hydroclimatic (precipitation, evapotranspiration (ET)) and anthropogenic (deforestation, land-use change...) changes are already threatening the region. Runoff changes in the basin have been characterized by more severe extreme values on the main stream for about forty years (Callede *et al* 2004, Espinoza *et al* 2009a) due to decadal modes of variability at regional scale (Marengo *et al* 1998, Labat *et al* 2004, Marengo 2004, Labat 2005, Espinoza *et al* 2006, 2009b) and to an interannual variability (Richey *et al* 1989, Marengo 1992, Guyot *et al* 1998, Uvo and Graham 1998, Ronchail *et al* 2005a, 2005b, 2006, Zeng *et al* 2008, Espinoza *et al* 2009a) leading to severe droughts such as the 1998, 2005 and 2010 events (Marengo *et al* 2008, Zeng *et al* 2008, Yoon and Zeng 2010, Espinoza *et al* 2011, Marengo *et al* 2011) and floods as experienced in 1999, 2009 (Chen *et al* 2010, Marengo *et al* 2010, 2012) and 2012. These events alter the livelihood of riverside populations (Drapeau *et al* 2011), the ecosystems and the vegetation functioning (Phillips *et al* 2009). Drought has been suggested to increase tree mortality and decrease photosynthetic activity in 2010 (Lewis *et al* 2011, Xu *et al* 2011). Furthermore the impact of such events on water resources (Bates *et al* 2008, Cox *et al* 2008), on fire and vegetation (Nepstad *et al* 2007, Cook *et al* 2012) might increase in the future. Are present-time extreme discharges precursors of future conditions?

Despite the uncertainties of future climate projections, related to CO_2 emission scenarios and to General and Regional Circulation Models (GCMs and RCMs respectively) (Li *et al* 2006, Vera *et al* 2006, Boulanger *et al* 2007, Marengo 2009), some consensus is found regarding climate change in the Amazon River basin. Indeed, a 4 °C warming is expected in tropical South America (Boulanger *et al* 2006) and a consensus is found about a precipitation decrease in austral winter (Vera *et al* 2006), a longer dry season and more severe droughts (Li *et al* 2006, 2008). Moreover, annual precipitation is expected to decrease in the eastern Amazon River basin and, in contrast, to increase and be more intense over the western Amazon (Meehl *et al* 2007, Alves and Marengo 2010).

Past studies on climate change and its impact on hydrology at global scale do not show any robust agreement about future runoff trends in the Amazon River basin. Some authors simulate a discharge reduction on the main stem of the Amazon River (Arora and Boer 2001) while others find a 5% increase in discharge at the end of the 21th century (Nohara *et al* 2006). Salati *et al* (2009), using 15 AR4 (4th Assessment Report) coupled models and the RCM HadRM3P (Hadley Centre Regional Model version 3P), project a 7–30% discharge decrease on the main stem of the Amazon River depending on the scenario (SRESB2 and SRESA2), the climate model and the period. At the regional scale, literature on future runoff projections in the Amazon River basin remains scarce because of hydrological modeling issues (Coe *et al* 2007, Decharme *et al* 2008, Beighley *et al* 2009, Trigg *et al* 2009, Getirana *et al* 2010, Paiva *et al* 2011a, Yamashima *et al* 2011, Guimberteau *et al* 2012, Paiva *et al* 2012). Milly *et al* (2005), using ensembles from 12 GCMs for the middle of the 21st century under the SRESA1B scenario, obtained a 5–10% runoff decrease in northeastern Amazon, contrasted to a 10–40% increase in western Amazon. Along the eastern edge of the Amazon River basin, Tomasella *et al* (2009) found a 30% (60%) mean annual (low-flow) decrease of the Rio Tocantins discharge at the end of the 21th century under the SRESA1B scenario. In the upper Amazon, Lavado Casimiro *et al* (2011) use two hydrological models, the climatic data from 3 GCMs under 2 SRES scenarios, and found both decreasing (4 basins) and increasing (3 basins) discharge trends in the Peruvian Amazon Andes basins. As a consequence of precipitation and runoff change, flooded areas and flood duration are found to increase in about one third of the basin, especially in Western Amazonia (Langerwisch *et al* 2012).

The above differences in future extreme flow estimations (low and high flows) for the Amazon River are due to the climatic forcing, model uncertainties but also to the lack of simulations at sub-basin scale. Will annual extreme flows be more severe during the 21st century than the present ones? Will there be differences within sub-basins? To try to answer these questions, our study focuses on sub-basin scale annual extreme values in the context of a changing climate. This was permitted by the availability of the Land Surface Model (LSM) ORCHIDEE (ORganising Carbon and Hydrology In Dynamic EcosystEms, section 2.1) which is able to accurately simulate the present-time streamflow in many stations over the basin (Guimberteau *et al* 2012). We build plausible scenarios of hydrological conditions that populations and ecosystems might have to cope with during the 21st century, based on a multi-scenarios approach from the IPCC's (Intergovernmental Panel on Climate Change) AR4 (Solomon *et al* 2007). Our methodology is first described (section 2). We detail the method based on anomalies to build the future climate forcings (section 2.2) and the hydrological signatures chosen to analyze the relationship between precipitation and streamflow (section 2.3). The hydrological signature results for present time are briefly given in section 3. The results from future simulations under SRESA1B at the middle of the 21st century are presented in section 4 and organized according to the different sub-basins. Section 5 summarizes the results obtained at the end of the century and under other scenarios.

2. Methodology

2.1. The land surface model ORCHIDEE

Hydrological simulations under climate change are performed using the hydrological module SECHIBA (Schématisation des EChanges Hydriques à l'Interface Biosphère–Atmosphère, Ducoudré *et al* 1993) of the LSM ORCHIDEE considering a 11-layer hydrology (De Rosnay *et al* 2002, D'Orgeval *et al* 2008, Campoy *et al* 2013) where a 2 m-soil is vertically discretized to calculate unsaturated soil water fluxes. The routing module (Polcher 2003, Guimberteau *et al* 2012) is activated in the model to simulate the daily transport of runoff and drainage to the ocean. Flooded areas are taken into account in ORCHIDEE using the representation of floodplains and swamps by D'Orgeval (2006) and a map of their spatial distribution over the Amazon River basin built by Guimberteau *et al* (2012). A detailed description of the LSM ORCHIDEE is given in the latter study.

2.2. Construction of climate change forcings

GCMs are known to poorly simulate precipitation patterns, even more at sub-basin scale (Solomon *et al* 2007). Moreover, because of the importance of local precipitation for hydrology, downscaling methods have been developed for hydrological studies and lead to improved results at basin scale (Wilby *et al* 1999). We applied the delta downscaling method approach to produce climate change forcings from GCMs results. This method is based on the addition of the monthly anomalies between two climatologies (simulated climate change and current climate results from a given GCM) to a baseline meteorological forcing. It appears to be a downscaling method well adapted for hydrological studies (Fowler *et al* 2007). In fact, if we assume that relationships between variables in the present-time climate are likely to be maintained in the future, the delta method incorporates from the present-time forcing more realistic spatio-temporal patterns in precipitation which are crucial for future streamflow simulation. However, this method prevents the study of future interannual variability as future monthly anomalies of climatologies are applied to the present-time forcing whose variability is maintained.

Climate change anomalies are derived from up to 8 AR4 GCMs (see table s1 available at stacks.iop.org/ERL/8/014035/mmedia), 2 different 20 yr running mean periods (2046–2065 and 2079–2098) and 3 emission scenarios (SRESB1, A1B and A2). We mainly focus on the results coming from SRESA1B, which projects a very rapid economic growth, a low population growth, a rapid introduction of new and more efficient technology, leading to a 720 ppm CO_2 concentration by the year 2100. This choice was driven by the better availability of the 8 atmospheric variables required to run ORCHIDEE (see table s2 available at stacks.iop.org/ERL/8/014035/mmedia) under SRESA1B scenario compared to SRESA2. The global meteorological dataset NCC (NCEP/NCAR Corrected by CRU data, Ngo-Duc *et al* 2005) is used as the baseline for this study. The 21 yr period representing the current climate in NCC dataset is the period 1980–2000, during which the precipitation dataset was corrected by ORE (Environmental Research Observatory) HYBAM (Geodynamical, hydrological and biogeochemical control of erosion/alteration and material transport in the Amazon River basin) (Cochonneau *et al* 2006) observations that have improved the present-time streamflow simulation (called 'ORCH4') by ORCHIDEE (Guimberteau *et al* 2012).

In each GCM grid-cell i, monthly anomalies ($\mathrm{ANO}(X_{\mathrm{GCM},i,t_{\mathrm{GCM}}})$) have been calculated from climatologies of the atmospheric variables outputs derived from each GCM ($X_{\mathrm{GCM},i,t_{\mathrm{GCM}}}$) (figure 1 and equation (1)). Linear spatial interpolation has been applied to the GCM outputs whose resolution was coarser than NCC (1.0°). The resulting monthly anomalies of a given variable of a GCM were applied to the corresponding NCC variable ($X_{\mathrm{NCC},i,t_{\mathrm{NCC}}}$) at each NCC time-step ($t_{\mathrm{NCC}} = 6$ h) for the corresponding month ($t_{\mathrm{GCM}} =$ 1 month) of each year of the 1980–2000 period (figure 1 and equation (2)). Thus, simulations are performed with ORCHIDEE forced by these new climate change forcings starting from the same state of equilibrium of the model (5 yr spin-up over the 1975–1979 period).

$$\mathrm{ANO}(X_{\mathrm{GCM},i,t_{\mathrm{GCM}}}) = \left(\frac{(X^{\mathrm{SRES}\alpha} - X^{20C3M})}{X^{20C3M}}\right)_{\mathrm{GCM},i,t_{\mathrm{GCM}}} \quad (1)$$

$$X_{\mathrm{NCC}_{\mathrm{futur}},i,t_{\mathrm{NCC}}} = X_{\mathrm{NCC},i,t_{\mathrm{NCC}}} + (\mathrm{ANO}(X_{\mathrm{GCM},i,t_{\mathrm{GCM}}}) * X_{\mathrm{NCC},i,t_{\mathrm{NCC}}}). \quad (2)$$

2.3. Hydrological signatures

The variations of three hydrological signatures (runoff coefficient, precipitation elasticity of streamflow and first/last deciles of streamflow) are studied at sub-basin scale using 16 ORE HYBAM gauge stations (table 1) to characterize the

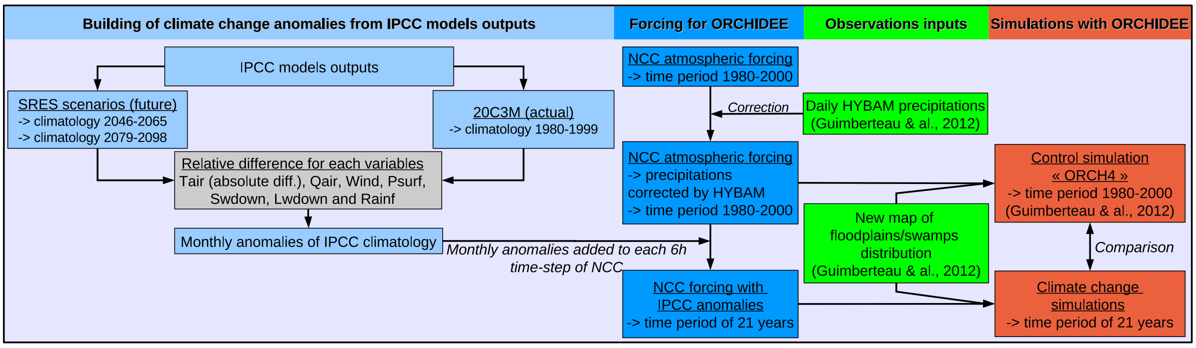

Figure 1. Flowchart of the future climate forcing construction.

Table 1. List of ORE HYBAM gauge stations over the Amazon River basin (see figure 2 for their localization on a map). For each station, mean annual streamflow (Q_{mean}), runoff coefficients (R_{coeff}) and streamflow elasticity to precipitation (ε_{p}) are computed from observations (ORE HYBAM database) and present-time simulation ORCH4, on average over the 1980–2000 period (except for the observed streamflow at the river mouth whose value is from Callede *et al* (2010), for the 1972–2003 period average). The colors indicate the localization of the stations in the basin (black: mainstream; red: southeastern/southern region; green: western region; blue: northwestern/northern/northeastern region).

Station	River	Lat	Lon	Area (km^2)	$Q_{\mathrm{mean}}^{\mathrm{OBS}}$ (m^3 s^{-1})	$Q_{\mathrm{mean}}^{\mathrm{ORCH4}}$ (m^3 s^{-1})	$R_{\mathrm{coeff}}^{\mathrm{OBS}}$	$R_{\mathrm{coeff}}^{\mathrm{ORCH4}}$	$\varepsilon_{\mathrm{p}}^{\mathrm{OBS}}$	$\varepsilon_{\mathrm{p}}^{\mathrm{ORCH4}}$
Mouth (AMAZ)	Amazonas	∼0.00	∼−50.5	5961 000	206 000	221 968	0.49	0.53	—	1.53
Óbidos (OBI)	Amazonas	−1.95	−55.30	4680 000	169 515	167 188	0.50	0.49	1.70	0.78
Manacapuru (MANA)	Solimoes	3.31	−60.61	2242 400	100 819	96 584	0.61	0.58	1.13	1.78
Fazenda Vista Alegre (FVA)	Madeira	−4.68	−60.03	1293 600	28 374	42 062	0.35	0.52	1.08	1.46
Porto Velho (PVE)	Madeira	−8.74	−63.92	954 400	19 418	27 923	0.34	0.49	1.36	1.02
Guajara-Mirim (GMIR)	Mamore	−10.99	−65.55	532 800	8 041	9 620	0.28	0.33	1.21	1.09
Rurrenabaque (RUR)	Beni	−14.55	−67.55	67 500	1 986	1 796	0.68	0.61	1.05	1.38
Altamira (ALT)	Xingu	−3.38	−52.14	469 100	8 000	15 327	0.29	0.55	1.43	1.74
Itaituba (ITA)	Tapajos	−4.24	−56.00	461 100	11 789	13 964	0.39	0.46	1.12	1.54
Sao Paulo de Olivença (SPO)	Solimoes	−3.45	−68.75	990 781	46 206	37 910	0.73	0.60	0.61	1.43
Tamshiyacu (TAM)	Amazonas	−4.00	−73.16	726 400	30 530	22 593	0.81	0.61	0.32	1.08
Labrea (LAB)	Purus	−7.25	−64.80	230 000	5 472	6 755	0.37	0.46	0.53	2.15
Gaviao (GAV)	Jurua	−4.84	−66.85	170 400	4 632	7 197	0.49	0.76	0.72	1.75
Acanaui (ACA)	Japura	−1.82	−66.60	251 800	14 075	17 633	0.60	0.75	0.46	1.10
Serrinha (SER)	Negro	−0.48	−64.83	291 100	16 193	16 354	0.59	0.59	1.10	1.26
Caracarai (CARA)	Branco	+1.83	−61.08	130 600	2 780	3 768	0.47	0.63	1.24	1.85
Sao Francisco (SFR)	Jari	−0.41	−52.33	51 343	990	2 024	0.39	0.80	1.22	1.68

hydrological response to future precipitation change. In the stations where it accounts, the backwater effect (Meade *et al* 1991) that is not represented in ORCHIDEE, is eliminated in the measurements (Espinoza *et al* 2009a). For each signature, differences are performed between one simulated year in climate change condition and the baseline ORCH4 climatology. Thus we estimate for each year the signature variation between future simulation and average present-time condition. The results are presented from an ensemble which is an average of 168 differences (8 models *21 yearly differences) between future and present conditions.

2.3.1. Annual runoff coefficient. The annual average runoff coefficient (R_{coeff}, equation (3)) varying from 0 to 1, is the proportion of the long-term average precipitation ($\bar{P}$ in kg m^{-2} s^{-1}) that runs off into streams (long-term average discharge $\bar{Q}$ in kg m^{-2} s^{-1}), assuming no net change in storage. A high (low) runoff coefficient points out that a large amount of water leaves the basin as streamflow (as ET).

$$R_{\mathrm{coeff}} = \frac{\bar{Q}}{\bar{P}}. \qquad (3)$$

Three runoff coefficients are computed for each sub-basin:

- $R_{\mathrm{coeff}}^{\mathrm{OBS}}$ is computed from ORE HYBAM discharge data (except for the river mouth discharge whose value is from Callede *et al* (2010), for the 1972–2003 period average) and from the precipitation dataset used to correct NCC forcing (see section 2.2), on average over the 1980–2000 period.

- $R_{\rm coeff}^{\rm ORCH4}$ is computed from simulated precipitation and streamflow values of the present-time simulation, ORCH4.
- $R_{\rm coeff}^{\rm ENSEMBLE}$, is computed from the averaged simulated precipitation and streamflow values of the different future simulations, for each scenario and time horizon.

2.3.2. Precipitation elasticity of streamflow. In order to quantify the response of discharge to the precipitation change, we compute the coefficient of elasticity ($\varepsilon_{\rm p}$, equation (4)) which is the relative change in mean annual streamflow divided by the relative change in mean precipitation (Schaake 1990, Sankarasubramanian *et al* 2001). The streamflow elasticity can vary usually between 1.0 and 4.0, i.e. a 10% change in mean annual precipitation results in a 10–40% change in mean annual runoff. The precipitation elasticity of streamflow is strongly correlated to the runoff coefficient, the mean annual precipitation and the streamflow according to Chiew (2006). Thus, river basins with low (high) runoff coefficient will be more (less) sensitive to precipitation change.

$$\varepsilon_{\rm p} = {\rm median}\left[\frac{(Q_y - \bar{Q})}{(P_y - \bar{P})} \cdot \frac{\bar{P}}{\bar{Q}}\right] \qquad (4)$$

P_y and Q_y (both in kg m^{-2} s^{-1}) are the annual precipitation and discharge respectively, averaged over the year y of the time period.

The potential evaporation elasticity of streamflow is computed in the same way according to equation (4).

2.3.3. First and last deciles of streamflow. The deciles are used to distinguish the monthly low flow (i.e. first decile) and monthly high flow (i.e. last decile) for a given sub-basin. The first (last) decile of streamflow is the value not exceeded by the lowest (highest) 10% of all streamflow values. They are calculated for each sub-basin, each year of future climate forcing and scenario, for present and future time horizons. For each future simulation, the significance of the median change of the first and last deciles is assessed using the Wilcoxon signed-rank test at the 95% level.

3. Brief overview of simulated streamflow related to precipitation in present time

The ability of ORCHIDEE to simulate the streamflow variations over the end of the 20th century was thoroughly tested over the Amazonian sub-basins by Guimberteau *et al* (2012). The streamflow simulation has been improved using the new ORE HYBAM precipitation forcing which better takes into account the precipitation regimes. They are very different according to the regions of the Amazon River basin due to its huge size and its extension in both hemispheres. We can distinguish different seasons of precipitation according to the regions (Figueroa and Nobre 1990, Nobre *et al* 1991, Ronchail *et al* 2002, Espinoza *et al* 2009a). From southeastern to southwestern regions, the precipitation season occurs in DJF (December, January and February) and the dry season in JJA (June, July and August). The precipitation seasonality in western regions is similar but less pronounced. The rainiest and driest seasons in northwestern regions are slightly more marked; they occur in MAM (March, April and May) and DJF respectively. The precipitation regime slightly differs in the northernmost and the northeastern regions where the rainiest seasons occur in MJJ (May, June and July) and MAM, respectively.

The hydrological signatures describing the link between precipitation and runoff (runoff coefficient and precipitation elasticity) in present time are briefly analyzed in sections 3.1 and 3.2, then used to assess future streamflow variations in a climate change perspective (section 4). In the latter section and section 5, we focus on the future change of the annual extreme values of streamflow using the hydrological signature of streamflow deciles.

3.1. Annual runoff coefficient

The runoff coefficient observed at the mouth ($R_{\rm coeff}^{\rm OBS} = 0.49$) (table 1) indicates that about half of the precipitation reaching the Amazon River basin runs off to the mouth. It is in good agreement with Callede *et al* (2008) who found values between 0.41 and 0.59 for the 1940–2003 period, with Shuttleworth (1988) who evaluated that one half of the incoming precipitation is returned to the atmosphere as ET above the Amazonian forest, and with Marengo's (2006) compilation of results indicating that no more than 46% of precipitation leaves the basin as runoff. The high $R_{\rm coeff}^{\rm OBS}$ values in the stations at the outlet of the Andean basins show the influence of the high precipitation values (hotspots) occurring in these regions and the topography that induces a fast runoff generation from soil layers toward the exit of the Andes. This is observed at Rurrenabaque (0.68) and Tamshiyacu where $R_{\rm coeff}^{\rm OBS} = 0.81$ is close to the value given by Espinoza (2009) (0.79). On the contrary, southern runoff coefficients varying from 0.28 (Guajara-Mirim) to 0.39 (Itaituba) are very low in regions retaining large amount of water in floodplains. Southern $R_{\rm coeff}^{\rm OBS}$ is in good agreement with Molinier (1992), Molinier *et al*'s (1995) estimations which give 0.36, 0.29 and 0.39 at Fazenda Vista Alegre, Altamira and Itaituba, respectively. Northern runoff coefficients generally exceed 0.5, with exception of the northernmost and tropical basins (Caracarai and Sao Francisco basins). They are overestimated compared to Molinier (1992) and Molinier *et al* (1995) (about 0.50 at Serrinha for the 1973–1989 period).

$R_{\rm coeff}^{\rm ORCH4}$ (table 1) sums up the streamflow results obtained in Guimberteau *et al* (2012), where the northern and western streamflow simulation was improved by ORE HYBAM precipitation forcing and new floodplains distribution in the model. The main western basins at Tamshiyacu and Sao Paulo de Olivença are subjected to a runoff coefficient underestimation, probably due to the lack of precipitation data in the northwestern Amazon (Azarderakhsh *et al* 2011). Sub-basins from southern regions present a large overestimation in runoff coefficient maybe due to an ET underestimation and an exaggerated extension of rainy spots in the forcing.

3.2. Precipitation elasticity of streamflow

The streamflow elasticity simulated in this study for present-time climate ($\varepsilon_{\mathrm{p}}^{\mathrm{ORCH4}}$) varies between 0.8 (at Óbidos) to 2.15 (Labrea) over the Amazon River basin and is generally overestimated compared to the observations ($\varepsilon_{\mathrm{p}}^{\mathrm{OBS}}$) (table 1). The streamflow elasticity computed over the Amazon River sub-basins is of the same order of magnitude as other estimations in the world: it ranges from 1.0 to 2.5 over the US (Sankarasubramanian *et al* 2001), from 2.0 to 3.5 in 219 catchments across Australia (Chiew 2006), from 1.0 to 4.0 in the Quarai River which is a tributary of the Uruguay River (part of the La Plata basin) (Paiva *et al* 2011a), from 2.0 to 6.0 for different LSMs in the Colorado River (Vano *et al* 2012) and from 1.0 to 3.0 at global scale according to Tang and Lettenmaier (2012).

4. How do the different Amazonian sub-basins respond to climate change (middle of the 21st century under SRESA1B scenario)?

We assess that low flow (high flow) is strongly related to the dry season (rainy season). This simple hypothesis is limited by the complex relation between precipitation and streamflow regimes within the Amazon River basin. Floodplains extension can modify the relationship between precipitation amount and extreme stream flows. However, for a small sub-basin such as the Mamore at Guaraja-Mirim, the lag between precipitation and runoff is not significant at monthly timescale. In contrast, it can be significant (about two months) over larger sub-basins including floodplains areas (Óbidos, Manacapuru, Fazenda Vista Alegre). After a brief overview of the climate change projected over the basin for the 2046–2065 time horizon under the SRESA1B scenario (section 4.1), the regional changes in extreme flows related to the seasonal precipitation changes (see section 3) are presented for three large coherent regions of the basin (see color code in table 1) in section 4.2: southeastern/southern region (SE–S, section 4.2.1), western region (W, section 4.2.2) and northern region including northwestern/northern/northeastern regions (NW–N–NE, section 4.2.3). A separate section is dedicated to Óbidos results (section 4.2.4), representing the hydrological behavior of the Amazon River basin.

4.1. Overview of projected climate change

Annual ensemble mean temperature increases by 2.04 °C (1.49 °C–2.63 °C according to the different simulations) on average over the basin. Higher annual increase occurs in southern sub-basins with the highest ensemble increase over the Mamore River basin (2.15 °C). The temperature increase leads to an enhanced ensemble mean ET by 7.0% on average over the basin. A slight precipitation increase by 1.1% is simulated over the basin. The precipitation change is spatially contrasted. At the annual timescale, few GCMs (generally 3 out of 8) simulate a precipitation increase (i.e. most GCMs project a precipitation decrease) in eastern (Xingu and Tapajos Rivers), northernmost (Branco River) and, to a lesser proportion, southern (Mamore and Beni Rivers) regions (figure 2(a)). In contrast, wetter conditions are found by a majority of GCMs (generally 7 out of 8) in western (Amazonas River at Tamshiyacu outlet, Purus and Jurua Rivers) and northwestern regions (Japura and Negro Rivers).

4.2. Regional response of streamflow to precipitation change

4.2.1. Southeastern/southern region. The annual precipitation decrease predicted by most models, associated to an ET increase (not shown), leads to a runoff coefficient decrease by more than 5% at the six stations of the region (figure 3). The persistent south-easterly precipitation decrease from March to November predicted by most GCMs (figures 2(c)–(e)) particularly affects streamflow at Altamira where a near 10% decrease in runoff coefficient occurs (figure 3). A south-easterly decrease in dry-season precipitation (figure 2(d)) leads to more pronounced low flows at all the stations of the region (figure 4(b)). The signal is rather robust in the region because at each station, 75% of the ensemble low-flow differences between future and present are negative. The effect is significant over the Madeira (Porto Velho and Fazenda Vista Alegre stations) and the Xingu (Altamira station) rivers where JJA precipitation decreases by 9% and 22% respectively and is associated with a 5% ET increase due to temperature increase (about 2.5 °C) over these basins (not shown). Up to 7 out of 8 simulations give a significant average decrease in the median low flow by about 30% (50%) at Porto Velho (Altamira). No change in high flow is simulated at southerly stations except at Altamira where 6 future simulations out of 8 give a significant decrease by 18% in average (figure 4(a)). Southern sub-basins tend to become more responsive to precipitation. Indeed, precipitation elasticity of streamflow increases by more than 15% at Guajara-Mirim, Altamira and Itaituba (figure 5). Potential evaporation elasticity of streamflow also increases in these stations and in Porto Velho (between 5 and 19%), except at Altamira where it decreases by 66% (not shown) suggesting that the Xingu River basin becomes more (less) responsive to precipitation (evaporation) change.

4.2.2. Western region. Most GCMs project an annual precipitation increase over all the western part of the Amazon River basin (figure 2(a)). However, annual runoff coefficients are little affected at Tamshiyacu and Sao Paulo de Olivença (less than 2% decrease), whereas they decrease by more than 5% at Gaviao and Labrea (figure 3), where the lower proportion in swamps and floodplains does not compensate for the change in precipitation contribution to runoff. The wetter projected conditions persist seasonally during the wet season (DJF), in particular in the northwesternmost part of the basin and along the eastern slopes of the Andes, where all the models simulate a precipitation increase (figure 2(b)). High-flow variation is not significant in the four westerly stations. At Tamshiyacu (upper Solimões), all the simulations show a median high-flow increase ranging from 5 to 25% (not shown) but only 3 simulations are significant (figure 4(a)). No significant low-flow decrease is simulated over the western

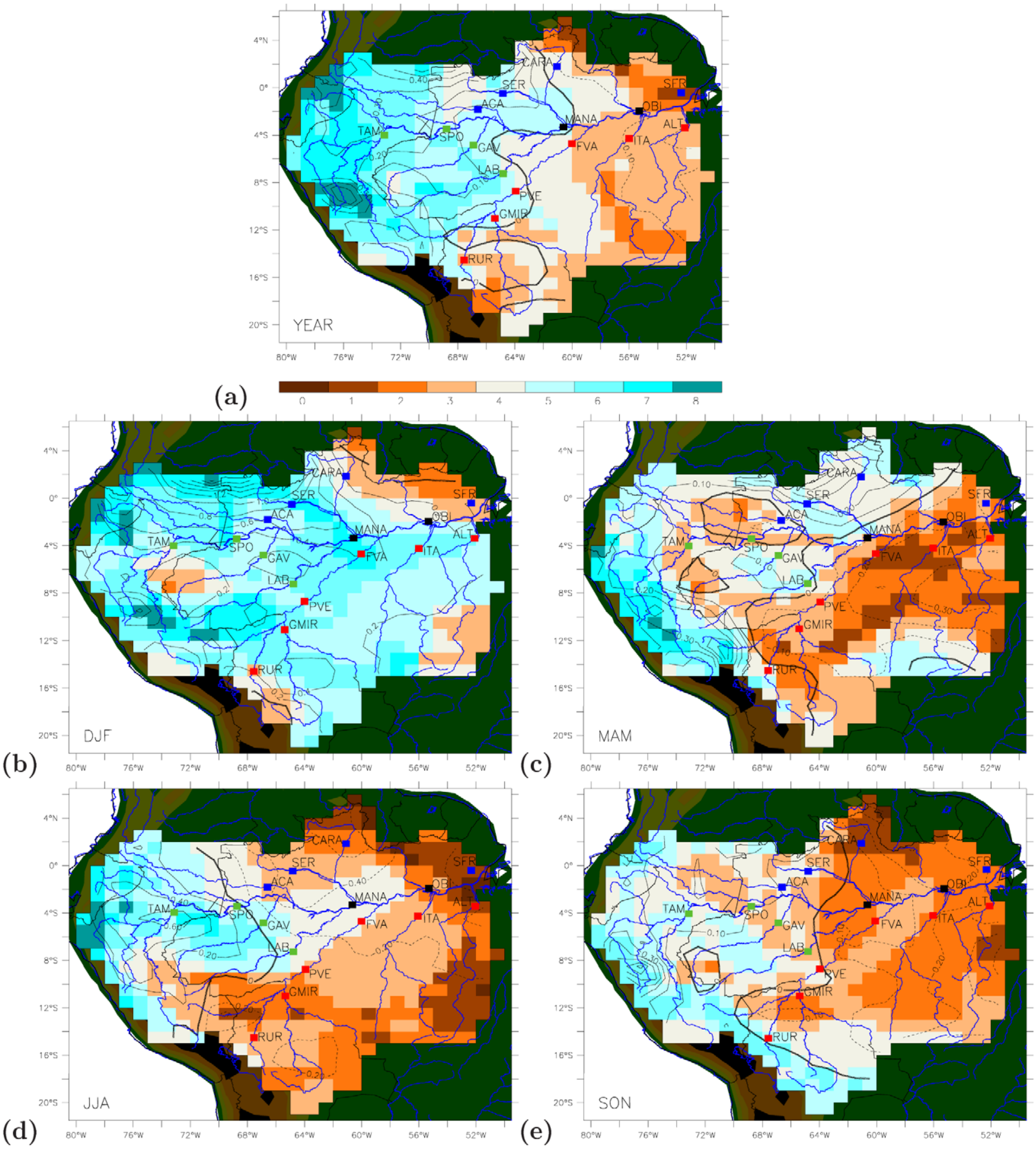

Figure 2. Number of GCMs out of eight that project a precipitation increase over the basin, for the 2046–2065 time horizon under SRESA1B scenario: (a) annual (b) DJF (c) MAM (d) JJA and (e) SON. Contour indicates the precipitation anomaly (mm d^{-1}) between ensemble and ORCH4 (solid line when the anomaly is positive and dashed line when negative). The localization of the ORE HYBAM gauge stations are indicated on each map with a color point (see table 1 for their coordinates and the color code).

part of the basin (figure 4(b)). Little change in precipitation elasticity of streamflow (less than 5% except at Labrea, figure 5) occurs at the westerly sub-basins, which include flooded regions that regulate the surplus of precipitation. These basins become more responsive to evaporation change where potential evaporation elasticity of streamflow increases up to 56% at Tamshiyacu (not shown).

4.2.3. Northern region. Most GCMs simulate an annual precipitation decrease in the north of the river mouth region and in the northernmost part of the basin (figure 2(a)). The tri-monthly decreases are projected with good confidence in the northeastern region except in MAM (figures 2(b)–(e)). Northerly streamflow decrease is going to be more severe from west (Acanaui) to east (Caracarai and Sao Francisco).

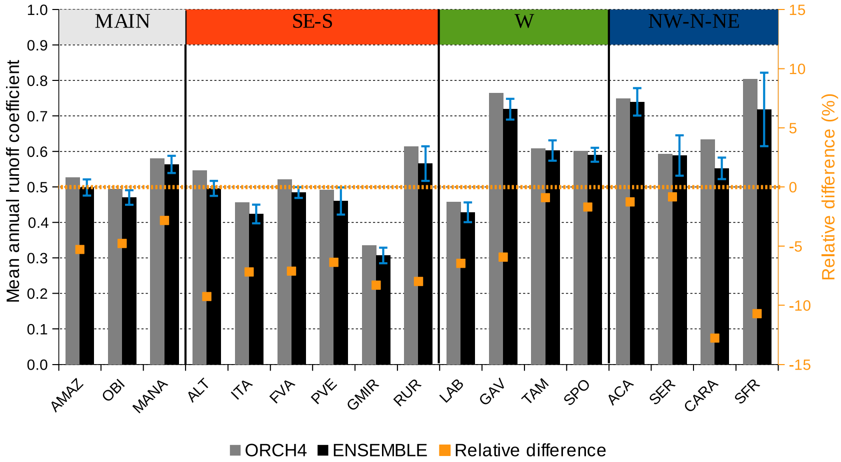

Figure 3. Mean annual runoff coefficients at the different stations across the Amazon River basin for ORCH4 simulation (in gray), ensemble simulation (in black, for the 2046–2065 time horizon under SRESA1B scenario) and mean annual relative difference (%) between both (points in orange). Blue lines on the ensemble histograms indicate the standard deviation. The colors on top indicate the localization of the station in the Amazon River basin (see table 1 for the coordinates of the stations and the color code).

Runoff coefficient highly decreases at Sao Francisco and Caracarai (10.5 and 13% decrease respectively) while low change (less than 2%) occurs at the most westerly stations (Acanaui and Serrinha) (figure 3). The same kind of severity is found regarding northerly low flows. They significantly decrease in at least six simulations out of eight at each station (figure 4(b)). Ensemble median low flows significantly decrease by 20% in the Japura (Acanaui station) and the Negro (Serrinha station) rivers, and it is much higher (55%) in the Branco River (Caracarai station) (figure 4(b)). The signal is rather robust for Serrinha and Caracarai, as 75% of the ensemble low-flow differences between future and present are negative. The discharge decreases at Caracarai during the high flow and recession periods. Overall, the seasonal variations of the ensemble are less contrasted than in the ORCH4 simulation (figure 6(a)). No significant change in ensemble high flow is simulated at Acanaui and Serrinha stations (figure 4(a)). Generally, the northern sub-basins at Acanaui and Serrinha are not responsive to future precipitation change (figure 5) but rather to evaporation; streamflow elasticity to potential evaporation increases by 22% in the Japura River basin (Acanaui station) (not shown). In contrast, the streamflow elasticity to precipitation (potential evaporation) increases (decreases) by 12% (22%) at Caracarai, suggesting that the Branco River basin becomes more (less) responsive to precipitation (evaporation) change.

4.2.4. Main stem of the Amazon River. The precipitation change is low on average over the basin, even though it is strongly spatially contrasted. At Óbidos station, the last gauged station before the mouth of the Amazon River, the runoff coefficient slightly decreases (about 4%) (figure 3) and no change in precipitation elasticity of streamflow is found (figure 5). The streamflow seasonal cycle only changes during the recession period when all the future simulations project more pronounced low flow than ORCH4 (figure 6(b)). Ensemble simulations provide a 10% median low-flow decrease (figure 4b) but only five future streamflow simulations out of eight give a significant decrease. No change in high flows is simulated at Óbidos station (figure 4(a)).

5. Hydrological extreme variations for the end of the 21st century and under other scenarios

The annual increase in mean ensemble temperature over the Amazon River basin does not depend much of the SRES scenario at the middle of the 21st century. But differences between scenarios are pointed out at the end of the century (+2.0, +3.0 and +3.8 °C for SRESB1, SRESA1B and SRESA2 respectively). The spatial contrast between an eastern decrease and a western increase in ensemble precipitation occurring during the middle of the century under SRESA1B is maintained for the end of the century. However, rainfall increase is accentuated in the west while the precipitation decline is confined in the north-east. In contrast, the eastern rainfall decrease is enhanced under SRESB1 at the end of the century. Regarding SRESA2, the eastern decrease is much weaker than under SRESA1B at the middle of the century and increased rainfall dominates at the end of the century, being much stronger in the west and in the north.

Considering the A1B scenario, the mean ensemble discharge simulations for the end of the century present a low-flow attenuation in western rivers (Madeira, Purus and Jurua) and on the main stem, when compared to the middle of the century. The most striking are observed at Gaviao and Labrea, on the Purus and Jurua Rivers, where the deficit falls to −5%, when compared to the present, instead of −30% at the middle of the century. In contrast, northwestern high-flow increase (Amazonas and Negro rivers) is enhanced (+12% at Tamshiyacu instead of +7% at the middle of the century) in agreement with the western rainfall increase at the end of the century.

No change in low flow under SRESB1 is found when compared to SRESA1B results for both time horizons. Eastern high flow decreases are slightly attenuated under SRESB1 when compared to SRESA1B at the middle and the end of the

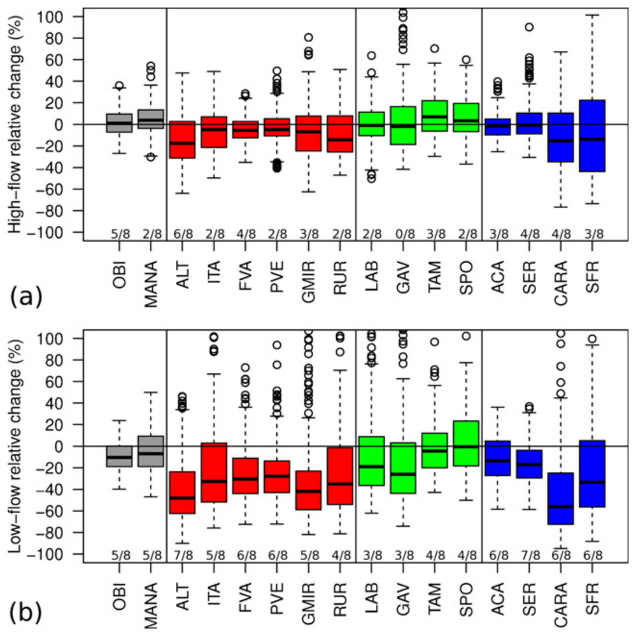

Figure 4. Relative change (%) of the (a) last deciles (i.e. high flow) and (b) first deciles (i.e. low flow) of streamflow between ensemble and ORCH4, for each station, for the 2046–2065 time horizon under SRESA1B scenario. For each year of future-time simulation, differences were performed between the result of the future-time simulation and the climatology of the present-time simulation. The boxes correspond to the interquartile range (IQR, the distance between the 25th and the 75th percentiles), the bold horizontal line in each box is the median and the whiskers extend from the minimum value to the maximum value unless the distance from the minimum (maximum) value to the first (third) quartile is more than 1.5 times the IQR. Circles indicate the outliers that are 1.5*IQR below (above) the 25th percentile (75th percentile). The colors of the boxes indicate the localization of the station in the Amazon River basin (see table 1 for the coordinates of the stations and the color code). The numbers of future simulations out of 8 that give a significant change of the median (Wilcoxon signed-rank test at the 95% level) are indicated below each box.

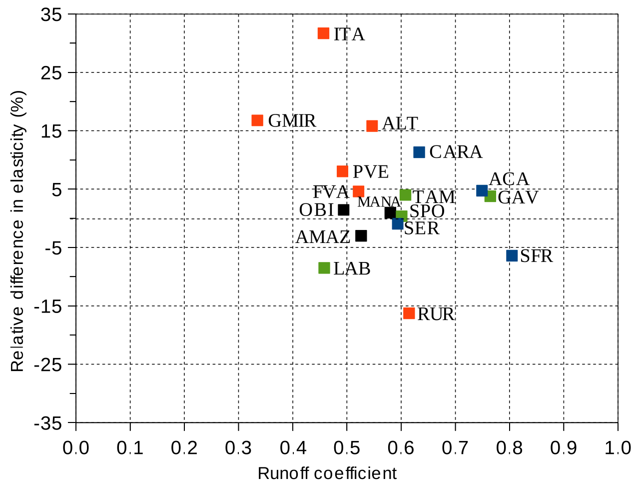

Figure 5. Scatter plot of relative difference (%) in streamflow elasticity to precipitation changes (between ensemble (2046–2065 time horizon under SRESA1B scenario) and ORCH4) and annual ORCH4 runoff coefficient. The colors of the squares indicate the localization of the station in the Amazon River basin (see table 1 for the coordinates of the stations and the color code).

century. Western high-flow increase under SRESB1 is lower than under SRESA1B at both time horizons. No change in high flow is observed at Óbidos.

The comparison of the results obtained under SRESA2 with those under other scenarios is biased by the fact that only 5 climatic simulations are available instead of 8. Nonetheless, the available data show that extreme discharge anomalies do not differ much from SRESA1B results in the middle of the century. By contrast, at the end of the century, and contrarily to the results obtained under B1 and A1B scenarios, the low flow increases consistently in a large western part of the basin (from +10 to +30%) and consequently on the main stem (+10% at Manacapuru, +5% at Óbidos). Elsewhere, the low-flow decrease is of the same order than for SRESA1B. Western and northern high flow increases substantially, in agreement with rainfall variation, leading to a 10% increase in Manacapuru and Óbidos. Interestingly for impact applications, intensification in extreme values (low and high flow) predominantly occurs in the Negro River at Serrinha (−20% and +20% respectively).

In conclusion, river discharge changes and differences between scenarios are less pronounced at the middle of the century and more accentuated at its end, especially between SRESA2 and the others; indeed, all year round, a large runoff increase is observed in a large part of the western Amazon River basin.

6. Conclusion

Future climate change should substantially modify the Amazon River basin hydrology and impact the water resources availability. In this study, we use the LSM ORCHIDEE to provide discharge projections, according to future climate forcings built from several GCM projections under the emission scenarios B1, A1B and A2. Various hydrological responses are found in the different sub-basins of the Amazon under SRESA1B scenario. Some of them are not significant since climate projections from the different GCMs diverge, as already assessed in previous studies (Bates *et al* 2008, Kay *et al* 2009, Blöschl and Montanari 2010, Nobrega *et al* 2011, Paiva *et al* 2011a), but in some sub-basins and for specific periods of the year, the projections are more reliable. In particular, low flows are projected to decrease severely in most stations, especially in the southern Madeira and Xingu Rivers and in the northern Branco River, and to a lesser extent in the northern Negro and Japura Rivers. By the middle of the 21st century, the decrease is projected to reach −50% (−55%) in the Xingu (Branco) Rivers. A low-flow decrease is projected in Óbidos, on the main stem of the Amazon River, which has already been experiencing an increasing number of droughts over the last twenty years. In contrast, the western and upper part of the Amazon are projected to undergo an annual precipitation increase so that high-flow discharge is expected to increase by 7% in the middle of the 21st century and even more by the end of the century. Therefore,

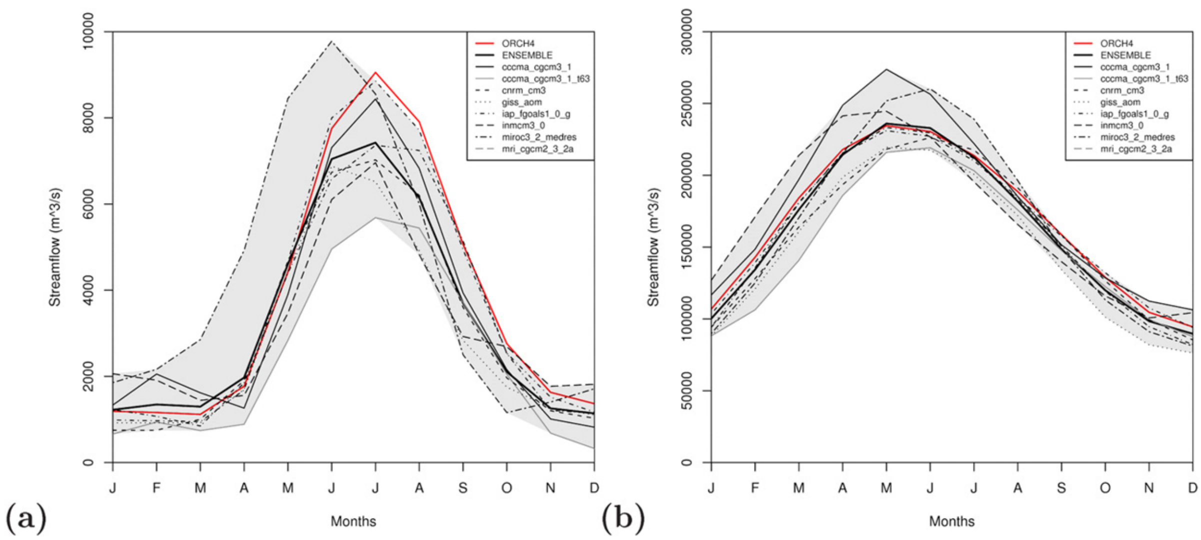

Figure 6. Mean seasonal streamflow (m^3 s^{-1}) simulated with eight simulations of future (gray lines) and ensemble (bold black line), for the 2046–2065 time horizon under SRESA1B scenario, at (a) station Caracarai (CARA) and (b) station Óbidos (OBI). The envelope (in gray) defines for each month the minimal and maximal values between the 8 simulations of future. ORCH4 (in red) is the present-time simulation for the present-time period 1980–2000.

more flooding events are expected in this region which is already affected by increasing extreme discharge. In the main stem of the Amazon River, however, there is no consistent signal of more flooding, neither of varying mean discharge, although previous studies projected a positive change (Nohara *et al* 2006) or a negative change (Milly *et al* 2005) of the Amazon runoff. By computing elasticity coefficients, we also show that the southern basins with low runoff coefficient are generally more responsive to precipitation change than the western basins with a high runoff coefficient. This result is in agreement with the findings by Chiew (2006) and Paiva *et al* (2011a).

Some uncertainties relative to the hydrological parameterizations of the LSM can be addressed. The present-time ET underestimation (Guimberteau *et al* 2012) and the infiltration parametrization are sources of uncertainties in future land water budget. Moreover, the routing module of ORCHIDEE does not represent the river dams and the backwater effect likely to affect the extreme stream flows in some tributaries of the Amazon (Meade *et al* 1991, Tomasella *et al* 2011). An advanced large-scale river routine module which represents backwater effect would represent more realistically the simulated water level in the rivers (Paiva *et al* 2011b, Yamazaki *et al* 2011, 2012). The modeling of vegetation growth with its interactive environment such as an interannual LAI variation and ET response to elevated CO_2, would lead to better realism using the STOMATE (Saclay-Toulouse-Orsay Model for the Analysis of Terrestrial Ecosystems, Viovy 1996) and LPJ (Lund-Postdam-Jena, Sitch *et al* 2003) modules. Other errors can result from the downscaling method. First, we assume that climate change varies only over large areas (i.e. as large as a GCM cell area) what leads to underestimate landscape heterogeneity effect due to the coarse resolution of the GCM. Thus, further simulations with better resolution than one degree would test if there is a dependence of simulated Amazonian hydrological processes on the spatial resolution of the forcing, as suggested by Verant *et al* (2004). Secondly, precipitation interannual variations in response to climate change could not be taken into account due to the simple downscaling method adopted in this study. Some bias-correction methods which conserve the changes of mean and standard deviation of the model simulation data (Watanabe *et al* 2012) could be adopted to preserve interannual variations. Uncertainties remain also in the emission scenario or the magnitude of rise in mean global temperature and the choice of the GCM. Indeed, a comparison between more AR4 GCMs and a new generation of GCMs (AR5) corroborates high consensus on drying during the dry season but highlights a lower confidence in wettening over the western part of the basin (Joetzjer *et al* 2013). Finally, considering land surface changes and deforestation in our LSM would probably affect the simulated mean and extreme flows over the Amazon River basin, since several studies reveal that land-use change accounts for at least 50% of the reconstructed global runoff trend over the last century (Piao *et al* 2007, Sterling *et al* 2012). The important deforestation in the Amazon River basin may in turn strongly alter the climate and eventually amplify the climate change (Malhi *et al* 2008). This topic, as well as improvement in the model parametrization, will be addressed in future work.

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